bonding exists between the metal atoms sharing the octahedral faces. Furthermore, these values indicate that the compounds with the composition $Li_xTi_{1.1}S_2$ ($0.1 < x \leq 0.3$) should be regarded as predominantly covalent compounds. In the corundum structure of Ti_2O_3, which can be regarded as an ionic compound, the Ti-Ti separation across the shared octahedral face is 2.58 Å (*4*).

A different phase exists for $Li_xTi_{1.1}S_2$ compounds with $0.5 < x < 1.5$. The x-ray powder patterns of these compounds do not resemble that of any titanium sulfide or that of $Li_xTi_{1.1}S_2$ ($0.1 < x \leq 0.3$). For the compound with $0.5 < x < 1.0$ the x-ray pattern can be indexed on a tetragonal cell with $a = 4.97$ Å and $c = 5.93$ Å. We did not find any superconductivity above 1.12°K for any of these compounds. However, we did discover an incomplete superconducting transition at approximately 2°K for the compounds with $1.0 < x < 1.5$. In this case the x-ray pattern is similar to that of the compound for which $0.5 < x < 1.0$, but it contains extra lines indicative of yet another slightly different compound.

It is difficult at the present time to explain the high transition temperature for the compounds $Li_xTi_{1.1}S_2$ ($0.1 < x \leq 0.3$). We can only state that, on the basis of the crystal chemistry of these compounds, they should not be regarded as intercalation compounds. Furthermore, in view of their superconductivity, these compounds do not seem to behave like intercalation compounds. The superconducting transition temperatures for this type of compounds range around 2°K. Recently Somoano and Rembaum (*5*) reported superconductivity in alkali-metal intercalation compounds of MoS_2. They found a superconducting transition temperature of approximately 1.3°K for Na_xMoS_2 and approximately 4.5°K for K_xMoS_2. After Rouxel *et al.* (*6*) reported the existence of K_xZrS_2, we found superconductivity, ranging from traces to a full bulk effect, in many other intercalation compounds, such as M_xZrS_2 and M_xHfS_2 with M = Na, K, Rb, or Cs. In these compounds the transition temperatures range between 1° and 3°K and are somewhat lower than in the corresponding MoS_2 phases. In almost all cases the transitions are partial and ill defined, and so no further effort was expended on these compounds. We have been unable to find superconductivity above 1.12°K in the corresponding compounds with TiS_2. The intercalation compounds were prepared by heating the constituents in a sealed silica tube to about 250°C and maintaining them at that temperature for several hours. It is thought that the compound $Li_xTi_{1.1}S_2$ with $x > 1.0$ is of an intercalated nature, because of its low transition temperature. But, if it is indeed an intercalation compound, it should be one of a new type of titanium sulfide.

In conclusion, we point out that these compounds constitute yet another example of high-temperature superconductivity at the expense of crystallographic stability (*7*). There is a large disparity in transition temperature between intercalation superconductors and three-dimensional systems. It is likely that many new superconductors with high transition temperatures exist, but these are not found because they are masked by their more stable and nonsuperconducting neighbors. This is the first time that such high transition temperatures have been observed in a hexagonal compound; all those that have been reported so far are cubic.

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Aftershocks Caused by Pore Fluid Flow?

Abstract. *Large shallow earthquakes can induce changes in the fluid pore pressure that are comparable to stress drops on faults. The subsequent redistribution of pore pressure as a result of fluid flow slowly decreases the strength of rock and may result in delayed fracture. The agreement between computed rates of decay and observed rates of aftershock activity suggests that this is an attractive mechanism for aftershocks.*

The frequency of occurrence of aftershocks following large shallow earthquakes decays with time. Typically the rate drops from thousands of aftershocks per day during the first day to 10 per day or less 100 days later. This decay requires a time-dependent process that is much faster than the large-scale tectonic loading and much slower than the propagation of elastic waves (*1*).

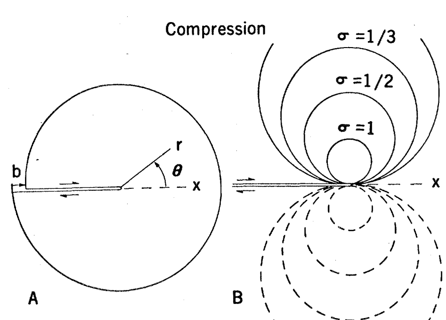

Fig. 1. (A) Representation of a fault's end by an edge dislocation with offset b; (B) the induced hydrostatic stress $\sigma(\mathbf{r})$. The initial pore pressure $P(\mathbf{r},0)$ is equal to $\sigma(\mathbf{r})$.

Theoretical and experimental investigations indicate that a viscous element is absolutely necessary in order to explain this time dependence of aftershock sequences. In model studies with friction and springs it has been impossible to produce aftershocks unless viscous dashpots are introduced. Similarly, computer models also require viscous elements to produce the necessary time-delayed event (*1, 2*). In other words, after a major earthquake there must exist some mechanism that "tells" a potential aftershock to wait before it occurs. Static fatigue is one possibility. However, its relevance to scales larger than laboratory samples has not yet been demonstrated. On the other hand, the decrease in the shear resistance of rocks in the laboratory with increasing fluid pore pressure has been well established, and its connection with earthquakes at Denver and Rangely, Colorado (3), induced by pumping fluids into wells, seems reasonably certain. Similar shocks associated with nuclear test explosions are probably also related to transient pore pressures and flow. Pore pressures in these cases were the result of the action of man. We decided to investigate the possibility (1, 4) that significant changes in pore pressure can also be generated by large earthquakes without the intervention of man. The decay of these pressure anomalies as a result of the flow of the pore fluid can provide the necessary viscous element in the form of an apparent time-dependent strength which can cause aftershocks.

When a fault breaks, it relieves part of the regional tectonic shear stress. However, because of inhomogeneities and particularly because of the finite length of the break, the strain during an earthquake involves substantial compression and dilatation as well as shear. If the rock in which these volumetric changes occur contains fluid (usually groundwater) in its pores, changes in pore pressure will also occur. These pressures are initially equal to the hydrostatic stress component (5) but will diminish with time as a result of the flow of the pore fluid.

The frictional strength of a rock is the shear stress above which sudden sliding or rupture takes place. Experiments show that, if a rock is subjected to a change in hydrostatic stress

$$\bar{\sigma} = (\sigma_{11} + \sigma_{22} + \sigma_{33})/3$$

and pore pressure P, the change in strength S is (6)

$$\Delta S = \mu_f(\bar{\sigma} - P)$$

where μ_f is the coefficient of friction. Immediately after an earthquake $\bar{\sigma} = P$ and the strength remains unchanged. As flow occurs, however, the stress-induced pore pressure approaches zero and ΔS consequently varies with time. The pore fluid flows from regions of compression to those of dilatation. This flow causes an increase of pore pressure and a consequent decrease in the strength in the original region of dilatation. Eventually, the strength at some point drops to the level of the local shear stress and an aftershock occurs. Since the main earthquake releases only a fraction of the regional tectonic load, the first motion or focal mechanism of the aftershock is likely to be similar to the main shock, in agreement with observations. Furthermore, since the aftershock releases only a very small fraction of the remaining regional shear stress, further increases in pore pressure should produce additional aftershocks.

We envision the aftershock sequence as analogous to the stick-slip process observed in laboratory experiments (7). If the confining and pore pressures of a sample (that is, the strength) are held fixed, and the applied increasing deformation causes the shear stress to equal the strength, a sudden slip occurs, associated with a sheer stress drop. Successive slip events occur as long as the deformation is continuously increased. The sheer stress rise, or drop, for each event is roughly constant. The same events may be produced by directly substituting increasing pore pressure for increasing shear stress. If the confining pressure is held fixed and the pore pressure is continuously and steadily increased (that is, decreasing strength), the stress drop will occur when the strength equals the shear stress. The increases in pore pressure between stress drops would be roughly constant. The number of possible shocks in a unit volume under shear stress would be approximately proportional to the total increase in the pore pressure in this volume, and the frequency (number of shocks per unit time) per unit volume would be proportional to the local time rate of change of the pore pressure. We thus expect the frequency of all aftershocks to be proportional to the integral of the time derivative of the pore pressure over the volume where the pore pressure is rising.

A very simple example illustrates our ideas. We approximate the end of the fault break by an edge dislocation in an otherwise homogeneous infinite elastic space. This is a vastly oversimplified model of a real earthquake. However, any break with an end or other inhomogeneity, whether it is abrupt, as in the dislocation, or more gradual, must have compression on one side of the crack and expansion on the other side.

In our model the hydrostatic stress induced by the sudden appearance of an edge dislocation (Fig. 1B) in a coordinate system r,θ about the end of the crack is (8)

$$\bar{\sigma}(r) = [\sigma_{rr}(r) + \sigma_{\theta\theta}(r)]/2 = A \sin\theta/r$$

where

$$A = \mu b/2\pi(1 - \nu)$$

b is the offset across the break, and μ and ν are, respectively, the shear modulus and Poisson's ratio for the material.

Typical values for μ and ν are about 10^5 bars and 0.25, respectively. At a distance of 1 km from the end of a fault with a 1-m offset, the change of shear and volumetric stress would be about 25 bars. The shear stress drop considered reasonable for large earthquakes and the fluid pumping pressure at Denver and Rangely are also of the order of 10 to 100 bars. It would therefore be rather surprising if the decay of the pore pressure field did not have profound effects on local seismicity. If, instead of an edge dislocation, we model the earthquake as a sudden stress drop on the surface of a crack, we find (9) that the magnitude of the compression or dilatation at the tips of the crack is also of the same magnitude as the stress drop. In fact, practically any local spatial variation of shear stress on the fault must be accompanied by a variation of compression and dilatation that is comparable in magnitude and size.

As we pointed out earlier, during the earthquake the pore pressure will change by an amount equal to the volumetric stress change. Thus Fig. 1B also represents the initial condition on the pore pressure for the fluid flow problem, namely,

$$P(\mathbf{r},0) = A \sin\theta/r$$

Later developments in the pore pressure field will be approximately governed (10) by the diffusion equation

$$\frac{\partial P(\mathbf{r},t)}{\partial t} = c\, \nabla^2 P(r,t) \qquad (1)$$

where c, the hydraulic diffusivity, is

$$c = K/\eta\beta$$

K is the permeability of the rock, η is the viscosity of the pore fluid, and β, the bulk compressibility of the system with porosity ϕ, is (11)

$$\beta = \phi\beta_{\text{water}} + (1 - \phi)\beta_{\text{rock}}$$

Because of the symmetry of the initial data, Eq. 1 is most easily solved by means of the Hankel transform. We find

$$P(\mathbf{r},t) = A\frac{1 - e^{-r^2/4ct}}{r}\sin\theta \qquad (2)$$

The direction of flow is primarily from the positive peak on one side of the fault to its negative image on the other side. The pressure at the peak decays as $t^{-1/2}$. However, because the peaks migrate slowly away from the fault, the pressure at a fixed point between the peak and the fault falls faster than $t^{-1/2}$. For large t, the pressure near the fault decreases as t^{-1}.

The time constant associated with the pore pressure decay is

$$\tau = L^2/4c$$

where L is a typical length. We take $L = 1$ km as a reasonable length. For unfractured solid rock, typical values for compressibility and permeability of $\beta = 5 \times 10^{-6}$ bar^{-1}, $K = 10^{-3}$ darcy yield $\tau = 120$ days. For fractured and jointed rock, $\beta = 5 \times 10^{-5}$ bar^{-1}, $K = 1$ darcy, and τ corresponds to 1 day. Thus τ will most likely be in the range from several hours to several months. This range of the time constant is similar to the periods observed in aftershock sequences, commonly lasting from a few days to several months.

We postulated earlier that the frequency of aftershocks in some volume near the main shock is proportional to the time derivative of the local pore pressure. Thus the total number of aftershocks per unit time is

$$\frac{dN}{dt} = \frac{1}{\alpha}\int \frac{\partial P}{\partial t}\, dv \qquad (3)$$

The proportionally constant α is equal to the increase in the pore pressure between successive fractures multiplied by some appropriate volume v. The integration is carried out over the volume where pore pressure is increasing, that is, the region where the main shock produces dilatation. For our two-dimensional dislocation model we need only integrate over the lower half-space in Fig. 1B. Differentiating Eq. 2 and substituting in Eq. 3, we find that the frequency of aftershocks

$$\frac{dN}{dt} = -\frac{1}{\alpha}\left(\frac{c\pi}{t}\right)^{1/2}$$

decays as $t^{-1/2}$.

The frequency of aftershocks for the 1966 Parkfield-Cholame, California, earthquake is shown versus time in Fig. 2. This sequence, one of the best ever recorded, shows an initial decay of $t^{-1/2}$ to $t^{-3/4}$ gradually shifting to t^{-1} at later time (*12*). Less detailed studies commonly show a decay of t^{-1} shifting to an exponential rate of decay after very long times (*13*).

The initial decay rate of the Parkfield-Cholame sequence is in remarkable agreement with our prediction. There are several good reasons, however, for expecting that our simple model gives only an approximate lower bound to the actual decay rate. First, the linear relation between frequency

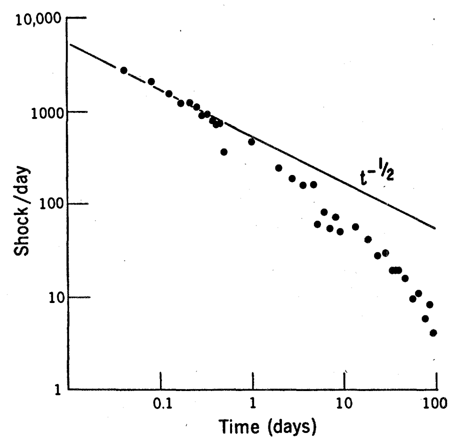

Fig. 2. The frequency of aftershocks versus time for the Parkfield-Cholame earthquake based on the data of Eaton (*12*). The solid line shows the slope predicted according to our model.

and the pressure-time derivative does not hold when the total possible number of aftershocks is small. In fact, if the possible pressure rise in a given volume is no longer large enough to overcome the threshold for a shock, the sequence in that volume ceases and it should be eliminated from the calculation of the integral (Eq. 3). It is the inability of the pressure field to overcome this threshold anywhere which prevents aftershock sequences for small earthquakes. Since the excluded volume clearly increases with time, the decay rate will also increase with time, as observed. Second, we have chosen a model which permits only two-dimensional pressure gradients. Any three-dimensionality would cause both the pore pressure field and the aftershock frequency to decay more rapidly. Finally, there are myriad inhomogeneities and nonlinearities that we have not considered. These include spatial variations of the various parameters, the pressured dependence of the permeability, and the coupling between the elastic and fluid flow problems referred to in (*10*).

We thus conclude that the flow of groundwater can provide the viscous element necessary to produce aftershock sequences. Changes in fluid pore pressure induced by earthquake faults are comparable with stress drops on faults and could therefore have a first-order effect on rock strength. The range of possible values of the time constant for the pore pressure decay and the rate of decay of the aftershock frequency are in general agreement with observations, although there is a clear need for more studies of the rate of decay during the first day after a major earthquake.

Our theory predicts clusters of aftershocks near the end of a fault. These are commonly observed. However, variations in stress release can produce volumetric changes and hence aftershocks anywhere along the fault. This mechanism therefore provides an attractive explanation for the aftershocks following large earthquakes, which are shallow enough to produce significant stress changes in regions where pore fluids are present. The lack of aftershocks following deep or small earthquakes is in agreement with our proposed mechanism. A critical test depends on pore pressure data associated with earthquake faults. Such a test must await continuous and extensive recording of water level and bottom pressures in existing oil, gas, and water wells in tectonically active regions.

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